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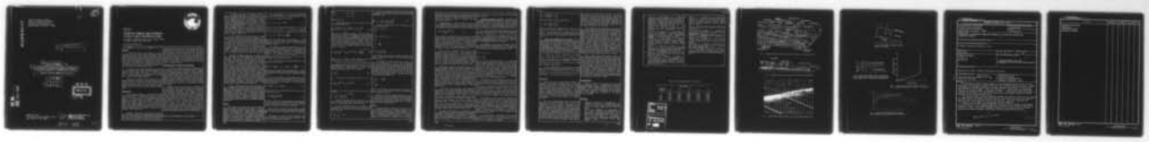

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EXPLANATION OF SUBMARINE LANDSLIDE MORPHOLOGY BY STABILITY ANALYSIS AND RHEOLOGICAL MODELS

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ABSTRACT

A theoretical study of mass movement on the Mississippi River delta front has been made using recently acquired field data and a simple rheological model. Recent measurements of sediment properties include cohesion, bulk density, pore pressure, and internal friction angle. Also, the geometry of a typical type of instability feature, an elongate slide, is examined in detail. A rheological model describing a Coulomb-viscous plastic in effective stress terms is proposed to describe certain features of the mass movement process.

The model is used to derive equations defining the initial failure of the slopes, the mass movement thickness and accelerations, the velocity of flow in the gullies, and the shape of the mud nose scarp in the toe area. The model indicates the importance of pore pressure in controlling sediment dynamics.

Results of the model compare favorably with the observed shapes of mud noses. Initial failures on low slopes (~0.5°) are attributed to pore pressures approaching geostatic values. Flow velocities are calculated to be several feet per second, based upon estimated sediment viscosities and pore pressures during flow.

INTRODUCTION

Recent detailed mapping of the Mississippi delta front has revealed the widespread occurrence of bottom features that appear to result from sediment mass movements.[1,2,3,4] The stability of these sediments has been generally discussed using the concept of effective stress;[5,6,7] however, there have been no attempts to analyze the dynamic aspects of the problem in this light. This report presents the results of a theoretical study that used recently acquired field data and a simple rheological model to examine the dynamics of mass movements on the delta front.

Although the geometry of the bottom features has been described, present understanding does not indicate the role of pore pressures in determining how movements take place, what the speeds or depth of flow may be, or how sediment properties may change during movement. Although direct measurements of these dynamic properties seem feasible,[8,9] they are not available at this time. Thus theoretical calculation of these values, until data become available, must be invoked. Also, the importance of a theoretical model as a tool to aid experimentation should not be neglected. Models define a set of specific parameters that would need to be measured and the accuracies that are needed. This information could greatly improve the efficiency of instrument design, data collection, and data analyses.

CHARACTERISTICS OF SUBMARINE MASS MOVEMENTS

A variety of slope and sediment instability morphologies have been identified and have been interpreted as collapse depressions, bottleneck slides, elongate slides, shallow rotational slumps, and mudflow gullies with overlapping depositional lobes.[4] A schematic diagram of the features, taken from Roberts et al.,[2] is shown in Fig. 1. Although each of the features possesses somewhat distinct surface morphologies, certain types appear to grade into others or to be linked spatially. This suggests that there may be a great deal in common among the physical factors that cause sediment failure and movement.

Examination of the various types of bottom features indicates that the most common and extensively occurring type is the elongate slide (Fig. 2). The slide has three basic components: (1) a somewhat circular area of multiple scarps and blocks, leading into (2), a long, linear gully, and terminating in (3), a composite toe fan or lobe area. The overall length of the slide may reach several thousand feet. The slump areas show curved scarps that vary from 10 to 25 ft in height and have slopes of 1° to 4°. The basal shear planes of the scarps are only slightly concave upward and tend to merge at depth into a single shear surface that is inclined parallel to the sediment surface. The average depth of the basal shear plane is less than about 100 ft, and its slope is about 0.5°. The slide, which begins as a shallow rotational slump, quickly becomes translational, showing substantial downslope displacements.

References and illustrations at end of paper.

The slump areas feed the elongate gullies or chutes, originally described as "delta-front valleys." Most of the gullies have widths ranging from 500 to 3,000 ft, averaging about 1,500 ft, and lengths up to 7 miles. Their surfaces are 6 to 25 ft lower than the adjacent bottom. They are bounded by sharp escarpments. Most of the gullies extend downslope approximately at right angles to depth contours, with slopes of from 0.2° to 0.8°, averaging about 0.5°. In plan view these are markedly sinuous, with alternating narrow constrictions and wider bulbous sections. The subsurface geometry of the gullies is not precisely known; however, the basal failure surface appears to be undulatory and not to exceed a depth of about 100 ft.

At the downslope end of the gullies are broad depositional lobes or toes, which are roughly circular in plan view. The edge of the depositional fan may be up to a few thousand feet from the gully mouth and possesses a distinct but rounded frontal scarp ranging in height from about 20 ft to as much as 70 ft, as shown in Fig. 3. Toe areas from different gullies may show two or three toe lobes, one on top of the other. The thickness of the displaced sediment in the toe area may be 100 ft; however, a single lobe appears to be closer to 50 ft thick. The general slope of the toe areas is 0.8° to 0.9°, or slightly greater than the chute or slump areas.

Although the sediment properties in a single elongate slide have not been systematically sampled, data have been taken in each component area at different locations. For example, a set of borings in a toe area has been taken in the South Pass area,[10] and the chemical, geologic, and geotechnical properties have been determined.[11,12] Borings were taken in front of a toe lobe, near the lobe scarp, and well back in the toe. The shear strength shows a weak surface layer about 60 ft thick having torvane strength values between 40 and 250 lb/sq ft with an average of about 100 lb/sq ft. At about the 60-ft depth, shear strengths increase rapidly to over 300 lb/sq ft. Water content of the sediment is generally high (>75%) in the surface layer but decreases abruptly to about 50% to 60% at the 70-ft depth. The upper layer also contains "severely disturbed sediment" with "fracture, slump or flow structure," as indicated in the geological analysis. Thus the toe lobes may be characterized as a sediment mass about 60-70 ft thick having a shear strength of about 100 lb/sq ft.

Values of drained cohesion and friction angles for delta sediments have been reported.[7] These measurements have not been specifically identified with either a particular depth below the mudline or as being from any component of a landslide. Thus the values may be taken as indicative of the scale size. The drained cohesion values range from 0 to 200 lb/sq ft, and friction angles vary between 20° and 26°. With this information and that concerning the landslide geometry, a rheological model can be quantitatively applied to delta-front sediment movements.

THEORY

The theory of plasticity has been successfully applied to glaciers[13] and subaerial mudflows.[14] Debris flows have been described[15] using several rheological models, including those for a simple plastic, a Bingham plastic, a Coulomb plastic, and Newtonian and non-Newtonian fluids. There are several similarities between mudflows, or debris flows, and submarine landslides. Therefore, the model used to describe subaerial mudflows[14] will be applied, but modified to include the effects of viscous forces, as suggested by Johnson.[15]

The shearing stress τ_s at a slope plane will be described by a cohesion C_s, an angle of internal friction ϕ_s, and "viscosity" M, such that

$$\tau_s = C_s + \sigma_N' \tan \phi_s + M(\dot{\varepsilon})^N \quad (1)$$

where σ_N' is the effective normal pressure, $\dot{\varepsilon}$ is the rate of shear strain, and $N \leq 1$. The angle ϕ_s is assumed small, so that $\tan \phi_s \cong \phi_s$. Equation (1) can be written

$$\tau_s = C_s - \gamma Z \left(1 + \frac{p}{\gamma Z}\right) \phi_s + M\left(\frac{du}{dZ}\right)^N \quad (2)$$

where γ is the bulk weight of the sediment, u is the horizontal velocity, Z is the depth (less than zero) measured from the mudline, and p is the pore-water pressure at depth Z.

The movement of submarine sediments will be considered as a two-dimensional problem. A cartesian coordinate system is used, the origin is at the mudline, Z is positive upward, and X is positive in the direction of movement. The sediments rest upon a rigid bottom at depth $Z = -h_b$, which slopes at an angle of β. It is assumed that β is small, so that $\sin \beta \cong \tan \beta \cong \beta$. The coordinate system is shown in Fig. 4.

The conservation of horizontal momentum for the sediments can be written

$$\rho \frac{du}{dt} = \gamma\beta + \frac{d}{dZ}\left[C_s - \gamma Z \left(1 + \frac{p}{\gamma Z}\right) \phi_s + M \left(\frac{du}{dZ}\right)^N\right] \quad (3)$$

where ρ is the density of the sediment and t is time. This equation balances downslope acceleration against the driving force of gravity and the retarding viscous forces.

First, the stability of the sediment can be considered as the case when the sediment is at rest and the gravitational force is less than the frictional resistance. The sediments are stable (i.e., do not accelerate downslope) if

$$-\gamma\beta \leq \frac{d}{dZ}\left(C_s - \gamma Z \left(1 + \frac{p}{\gamma Z}\right) \phi_s\right) \quad (4)$$

If equation (4) is integrated vertically (from Z to 0) it can be rewritten as

$$1 \leq \frac{C_s - \gamma Z (1 + p/\gamma Z)\phi_s}{-Z \gamma\beta} \quad (5)$$

This form is identical to the equation for the stability of planar slides on infinite slopes.[16] Thus equation (5) is a criterion for failure (or acceleration of the sediment downslope) when equality is achieved.

Next consider the case of steady flow when, sometime after failure, the sediments are moving at constant speed. In this case the term du/dt vanishes in equation (3).

The equation can then be used to determine the velocity profile, u = u(Z), within the moving mass. We specify the boundary conditions that

$$\frac{du}{dZ} = 0 \; . \; . \; . \; . \; . \; .\text{at } Z = -H_o$$

$$u = 0. \; . \; . \; . \; . \; . \; . \; .\text{at } Z = -h_b \; . \; . \; . \; . \; . \; . \; . \; .$$

where $-H_o > -h_b$.

Integrating equation (3) with respect to Z and rearranging terms gives

$$\frac{du}{dZ} = \left[\frac{-\gamma Z(\beta - \phi_s (1 + p/\gamma Z)) - C_s + A}{M}\right]^{1/N} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .(6)$$

where A is an integration constant. This is an equation from which $u = u(Z)$ can be determined if the bracketed term can be integrated. The assumption is made that the pore pressure p increases approximately linearly with depth, so that the term $p/\gamma Z$ is a constant, called G. This assumption appears to be valid within a few percent. Now the bracketed term in equation (6) is a constant with Z and can be integrated. After integration and use of the boundary condition, the velocity profile is given by

$$u = \frac{N}{N+1}\left(\frac{\gamma\varepsilon}{M}\right)^{1/N} (h_b - H_o)^{(N+1)/N}\left[1 - \left|\frac{-Z-H_o}{h_b - H_o}\right|^{(N+1)/N}\right] .(7)$$

where $\varepsilon = \beta - \phi_s(1-G)$. The variable ε might be considered the "effective slope" of the bottom.

The depth $-H_o$ is the point below the mudline where the gravity shear forces first equal the mud resistance and represents the lower boundary of a plug of nonsheared sediment. From equation (5), setting $Z = -H_o$, and taking the equal sign, it is found that

$$H_o = \frac{C_s}{\varepsilon\gamma} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .(8)$$

The velocity of the plug U_p is given by u at $Z = -H_o$

$$U_p = \frac{N}{N+1} \frac{\gamma\varepsilon}{M}^{1/N} (H_o - h_b)^{(N+1)/N} \; . \; . \; . \; . \; . \; . \; . \; . \; (9)$$

Therefore, the velocity profile below the plug can be written

$$u = U_p\left[1 - \left|\frac{H_o + Z}{H_o - h_b}\right|^{(N+1)/N}\right] . \; . \; . \; . \; . \; . \; . \; . \; .(10)$$

Therefore, the same model that describes the stability of the sediments has now yielded forms for the velocity profile under steady flow conditions.

We finally consider the case of sediment flow acceleration. Rather than consider the details of the accelerated flow, consider the vertically averaged horizontal velocity

$$\bar{U} = \frac{1}{h_b}\int_{-h_b}^{o} u \, dZ. \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

Taking the vertical average of equation (3) and assuming the bottom velocity shear can be approximated by $2\bar{U}/(H_o - h_b)$

$$\frac{d\bar{U}}{dt} = g\varepsilon - \frac{C_s}{\rho h_b} - \frac{M}{h_b}\left(\frac{2\bar{U}}{h_b - H_o}\right)^N \; . \; . \; . \; . \; . \; . \; .(11)$$

Taking the case for N = 1 and integrating with respect to time gives

$$\bar{U} = \frac{A}{B}(1 - e^{-Bt}) \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

and

$$\bar{X} = \frac{A}{B}\left(t + \frac{1}{B}(e^{-Bt} - 1)\right) \; . \; . \; . \; . \; . \; . \; . \; .$$

where

$$A = g\varepsilon - \frac{C_s}{\rho h_b} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

and

$$B = \frac{M}{\rho}\frac{2}{h_b(h_b - H_o)} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

as the speed of the flow and the distance traveled after time t. The average velocity U tends to a maximum value A/B with time.

The movement of mudflows has been described by Brüchl and Scheidegger.[14] A similar form of equation (1), without the viscosity term, was used to describe the soil properties. The leading edges of mudflows were described, assuming them to be at critical equilibrium. The equations for the shapes of the toe areas of the submarine landslides may also be given by the results. The curves for the toes are given in Fig. 5 for both active and passive Rankine states. The curves are given in dimensionless variables of

$$\eta = \frac{h}{H_o} \; , \; \xi = \frac{\varepsilon X}{H_o} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

and

$$A = \frac{2r\varepsilon C}{C_s} \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; . \; .$$

where h is the height of the mudline about the depth $Z = -H_o$, r has the value of about 1.8, and C is the cohesion of the unsheared sediment.

Therefore, using the Coulomb viscous model, a set of theoretical relations describing sediment movement and mudflow shape have been developed. These equations can be applied to the problem of landslides on the delta front by using the presently available data on sediment properties and landslide geometry.

RESULTS

A detailed comparison between the theoretical results and observations is not possible because of the lack of direct measurements of landslide dynamics. No direct observations of movement rates, depth of movement, or movement frequency are available. While several of the important soil properties (e.g., drained cohesion, pore pressure, viscosity, etc.) have been determined for delta sediments, these measurements have not been coordinated or systematically taken. With this lack of a proper data set, the application of the theoretical results to submarine landslides must be considered exploratory.

Initial failure. From the geometry of the features that have been identified, an initial analysis of the mechanisms of these subaqueous slope instabilities can be made in terms of equation (5), which at failure is

$$1 = \frac{C_s - \gamma Z\,(1 + p/\gamma Z)\,\phi_s}{-Z\gamma\beta} \qquad Z < 0 \;\ldots\ldots$$

It is possible to calculate from this equation the pore-water pressure (p) needed to initiate failure. The general slope angle for the delta-front features ranges from 0.1° to 1.1°, and the unit weight of the delta-front clays is approximately 100 lb/ft^3.[7] Calculations of pore-water pressure needed for failure have been made using various values of C_s and β and for a thickness of sediment of 50 ft. This depth represents a reasonable estimate of the depth of the basal shear plane.

Table 1 shows calculated values of pore-water pressure needed for failure. The pressures are expressed as the constant G or the pore-water pressure ratios to geostatic pressure $p/\gamma Z$. For all cases, the pore-water pressure needs to be very large for failure to occur. It must be in excess of hydrostatic pressure, and this constitutes strong artesian pressure. In addition, it is clear that the calculated values approach (or very slightly exceed) geostatic pressures, representing a condition of almost zero effective stress.

Preliminary results of pore-water pressure measurements within the subaqueous sediments on the Mississippi delta-front slope have been reported.[9] The piezometers were installed in block 28, South Pass area, at depths of about 26 ft and 49 ft below the mudline. The pore-water pressure data revealed large excess pressures after 7 hours of stabilization of the system, and values were considerably in excess of hydrostatic pressure. Significantly, the ratios of $p/\gamma Z$ approach the geostatic condition, with values of 0.895-0.946. Even larger pressures have also been recorded elsewhere in the offshore delta region, with $p/\gamma Z$ ratios of 0.986 at a depth of 50 ft within the sediment (Bennet, personal communication).

Thus, direct evidence indicates that large pressures occur within the delta-front slope sediments. Comparison of the measured pore-water pressure ratios ($p/\gamma Z$) with the calculated ratios needed for failure shows very close agreement. Indeed, using $C_s = 0$, $\phi_s =$ 20° (Henkel) and the highest ratios actually recorded, failure can be achieved. With larger C_s values, only relatively small increases in pore-water pressure are needed for failure. For example, for C_s = 200 lb/sq ft and β = 0.5° the pore pressure has to be increased by about 500 lb/sq ft for failure to occur. Fluctuations of 150 to 300 lb/sq ft have been recorded resulting from the passage of waves (Bennet, personal communication).

The spatial distribution of the landslides[4] seems to reflect the fact that primary failures are generally localized in shallow water or downslope of the depositional lobes of landslides in shallower water. Furthermore, the pore pressures need to be at or exceeding geostatic pressures. Pore pressures after failure may then be expected to increase further, exceeding geostatic values.

Flowage of sediment in gullies. The behavior of sediments in the chute areas can also be addressed using theoretical results. Given that the excess pore pressures which built up and exceeded the failure criteria are maintained, then the average velocity of the flows can be calculated. Initially sediment speed would increase until quasi-equilibrium condition was reached.

The velocity change during acceleration can be determined from equation (11). Taking as an example β = 0.5°, C_s = 10.0 lb/sq ft, h_b = 50 ft, ϕ_s = 20°, and G = 1.0 , the acceleration has a maximum value of .216 ft/sec^2. After only 90 sec of acceleration the velocity reaches a maximum value of 6.5 ft/sec. These velocities are highly dependent upon the pore pressures. When G = .99 , the maximum velocity is 2.4 ft/sec.

The parameter ε will play an important role in determining the flow characteristics of the sediments. It is a function of the bottom slope, friction angle, and pore pressure ratio to geostatic pressure. Thus

$$\varepsilon = \beta - \phi_s\,(1 - G) \;\ldots\ldots\ldots\ldots\ldots\ldots$$

where $G = \frac{p}{\gamma Z}$

The lower limit of the range of values of ε is determined by the failure criterion. From Table 1, the ratio G at failure had values from 0.98 to about 1.14, for β = 0.50 and ϕ = 20°. Therefore, ε has minimum values ranging from about 0.001 to 0.06. Thus, during flow, values greater than these are expected. The upper limit of ε obtainable is not easy to define; however, for a pore pressure ratio to geostatic pressure of 1.25 ε has the value 0.100. Therefore, the range of variation of ε may be taken as between about 0.0 and 0.1. Once the slope and the friction angle ϕ_s are fixed, ε is directly proportional to the pore pressure ratio G.

The properties of the sediment flow within the chute areas can be estimated from equations (8), (9), and (10). Assume that the chute sediments can be characterized by β = 0.5°, C_s =10.0 lb/sq ft, ϕ_s = 20°, h_b = 50 ft, and γ = 100 lb/cu ft. The thickness of the plug or rafted material, from equation (8), is 33.3 ft for ε = .003 and 40 ft for ε = .0025. Thus most of the sediment column within the chute would be unsheared, with the shear zone concentrated in the bottom 10 ft to 17 ft. If the point is reached where ε decreases to its minimum value, in this example .002, then H_o = 50 ft.

The velocity of the plug can be estimated if values of M and N are known. The values of M and N are not well known for delta sediments. General estimates for a Newtonian viscosity (i.e., N = 1) have been set at 15 to 180 lb-sec/sq ft. The viscosity of subaerial debris flows has been estimated for a Bingham-Newtonian viscous model at 80,000 to 450,000 times that of water, or viscosities 3 to 15 lb-sec/sq ft.[15] Estimates, however, of non-Newtonian viscoelastic properties of clays indicate N of about 0.07, which indicates strong nonlinear behavior.[18]

As an illustration of the method for determining the flow velocity profile, consider the case for N = 1 and M = 10 lb-sec/sq ft. The Newtonian model for the plug velocity, equation (9), becomes

$$U_p = \frac{1}{2}\left(\frac{\gamma\epsilon}{M}\right)\left(H_o - h_b\right)^2$$

and the veolcity profile, equation (10),

$$u = U_p\left[1 - \left(\frac{H_o + Z}{H_o - h_b}\right)^2\right]$$

For the assumed chute characteristics with G = .99 , then U_p = 4.2 ft/sec; and for G = 1.0 , then U_p = 6.5 ft/sec. The velocity profile is shown for G = .99 in Fig. 6. The flow velocity is inversely proportional to the viscosity M, so that doubling M reduces U_p to one-half.

Flow of the sediments in the toe area. The toe areas provide the first opportunity to check the results of the theory with actual data. This data concerns the shape of the toe noses using the theory of Brüchl and Scheidegger[14] for comparison. The toe noses display a curving edge that is shown in Fig. 3. The first example is the mudflow toe near South Pass area. A comparison between the observed toe shape and the best fit A $\cong$ 0.3 to the theoretical profile is shown in Fig. 7. The values used in the example were h_b = 60 ft, γ = 90 lb/cu ft, β = 0.021, and ϕ_s = 20°. The best fit of the curve was ϵ = 0.022, which implies C/C_s = 3.8 and C_s = 119 lb/sq ft. This value of ϵ is consistent with the value at failure, and the C_s value is reasonable. The calculated ratio of C/C_s implies C of 452 lb/sq ft, a value that is higher than expected.

A second example was considered from the Southwest Pass area. Selected points are shown in Fig. 7. The data is for ϵ = 0.02 and h_b = 40 ft. The values indicate C_s = 72 lb/sq ft and a C/C_s ratio of 2.8, or C = 202 lb/sq ft. The best curve would be about A $\cong$ 0.2.

DISCUSSION

The results of this study seem to support the application of a simple rheological model to the mass movement of sediments on the Mississippi delta front. The rheological model used combines a static Coulomb friction stress with a viscous stress, much in the manner used to describe subaerial mudflows. The data that is available (mud nose shape) seems to indicate general agreement with theoretical predictions. Where data is not available, the predictions using the theory (velocity, accelerations, pore pressures, etc.) seem to be reasonable, or at least correct to an order-of-magnitude.

The results indicate that flow in the toe area takes place in an active Rankine state, or when the sediments are subject to tension. This state is also indicated in the slump areas by the presence of tensional cracks and similar features. Within the gullies passive conditions may develop at points where gully depth or width causes decelerations and compressions of the sediments. Transportation of large blocks of sediment can be explained because the upper layer of the flow is unsheared and acts as a rigid surface. The combination of the rigid upper layer and a sheared lower layer has been described as plug flow, and this would seem to be the case for the offshore landslides. Shear zones at the sides of the gullies could be expected. The thicknesses of the plug and the shear zone are functions of the bottom slope and the pore pressures within the sediments. As these change during downslope movement, the thickness of the layers could change. Apparently in the toe area the sheared layer decreases in thickness as the flow decelerates, so that when stopped the toe thickness reflects the plug thickness.

As encouraging as the results may be, they must be evaluated in light of the present general lack of knowledge concerning deformation dynamics. The rheological model used is not the only one that could have been applied, and through back calculation ("fudging"), be forced to produce reasonable results. Even with the model used, slight changes in parameters (i.e., viscosity M or power N) that are poorly known would greatly change the predictions, making them at times indefensible. Another limitation to these results stems from the assumptions employed in applying the model, i.e., steady states, constant pore pressure, constant cohesion, etc. When applying this approach to a specific site, probably any specific site, these assumptions will almost certainly be violated. Finally, the data used in this study have been taken from several sources or are outright estimations (values for M and N), so that a verification of the predictions cannot be claimed.

These results indicate that further studies of submarine mass movements should consider the coordination of descriptive studies of feature geometries (slump areas, gullies, toe areas), selection and placement of in situ instruments (spatially and with depth), and determination of geotechnical (rheological) properties of the sediments. One of the most interesting and potentially important prospects for future study is that mass movements, as exemplified by an elongate slide, may be a sequence of interrelated events separated in space by thousands of feet, extending from shallow to deep water and separated in time by hours, if not days. This possibility means that the value of making measurements at a single site to explain or predict sediment movement at that same site may be quite limited.

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Table 1. Pore Pressure Ratios G at Failure

Cohesion lb/sq ft	Slope Angle (°) 0.3	0.5	0.7	0.9	1.1
0	0.985	0.976	0.966	0.956	0.946
50	1.013	1.003	0.994	0.984	0.974
100	1.041	1.031	1.021	1.012	1.002
200	1.096	1.087	1.077	1.067	1.058
300	1.152	1.142	1.133	1.123	1.113

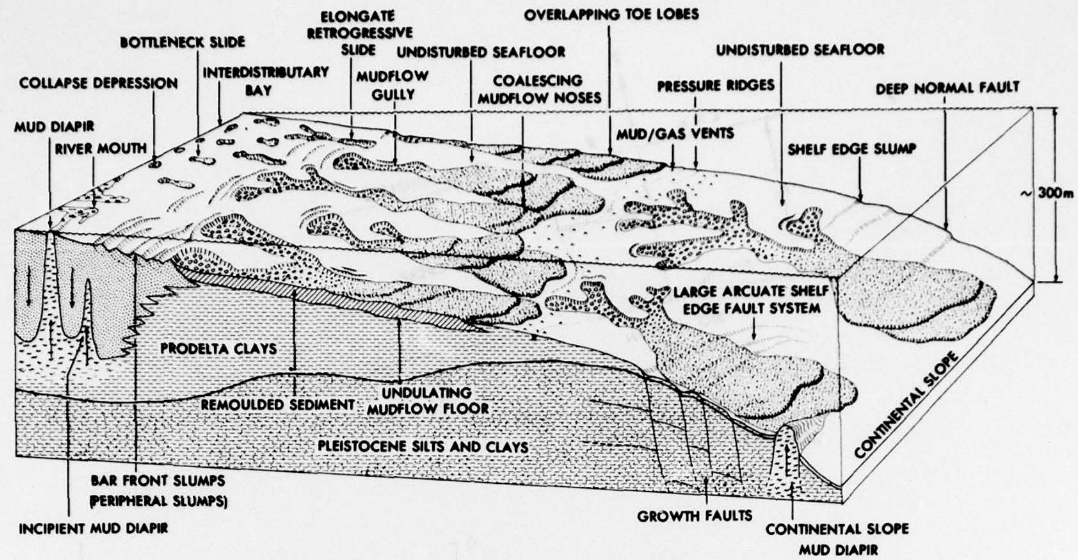

Fig. 1 - Schematic diagram of delta front instability features. (From Roberts et al.[2]).

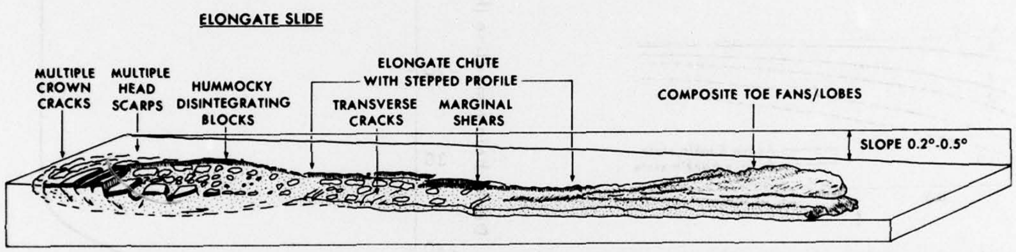

Fig. 2 - Schematic diagram of an elongate slide. (From Prior and Coleman[3]).

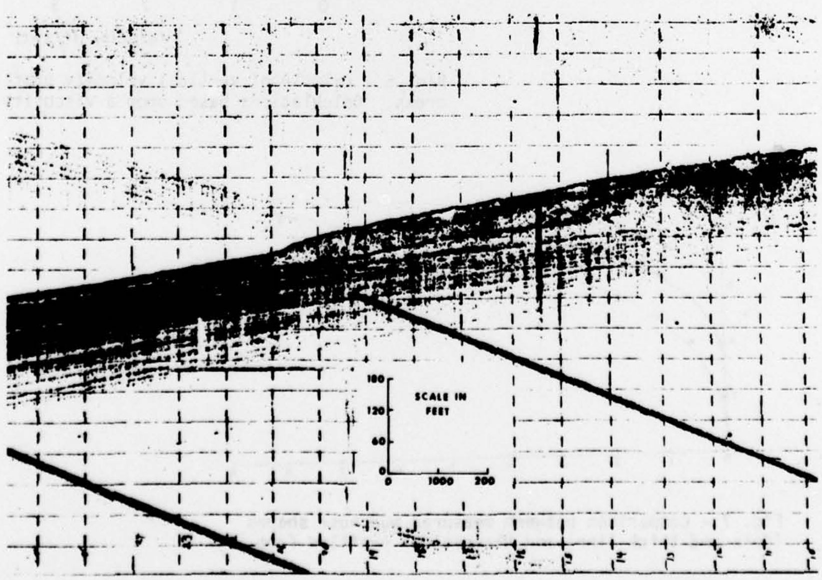

Fig. 3 - Sparker record showing depositional toe and the rounded frontal scarp.

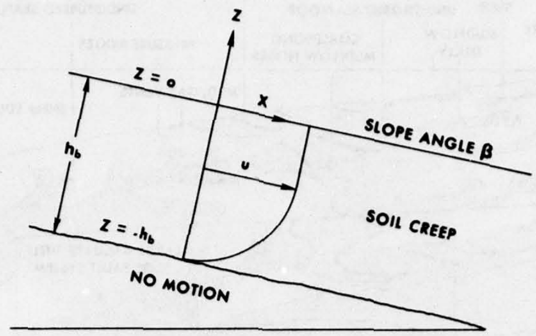

Fig. 4 - Illustration of coordinate axis and geometry of soil model.

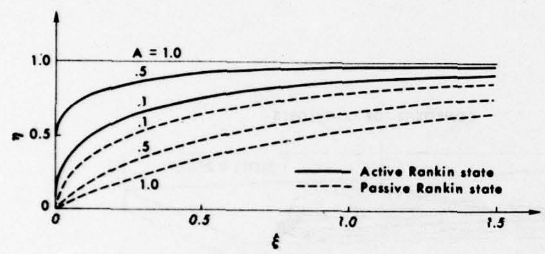

Fig. 5 - Theoretical profiles of earth flow tongues for various values of the parameter A and in both active and passive Rankine states. (From Bruchl and Scheidegger[14]).

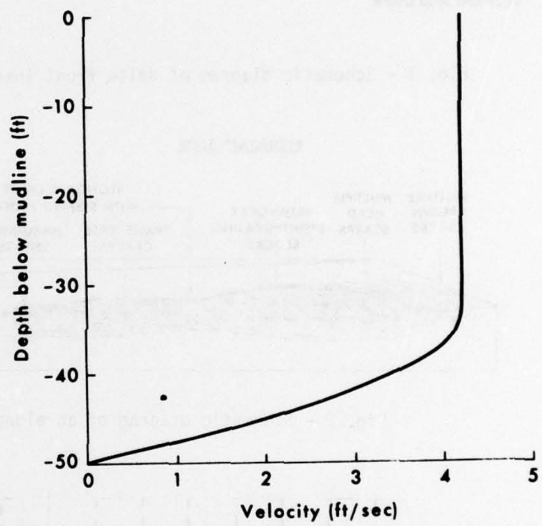

Fig. 6 - Example of vertical velocity profile in the gully areas. Calculations based upon a viscosity of 10 lb-sec/sq ft.

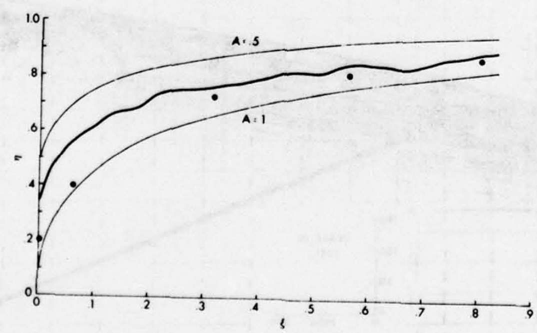

Fig. 7 - Comparison between measured mud nose shapes (dots and thick line) and theoretical profiles from Fig. 5.

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ABSTRACT

13. ABSTRACT

A theoretical study of mass movement on the Mississippi River delta front has been made using recently acquired field data and a simple rheological model. Recent measurements of sediment properties include cohesion, bulk density, pore pressure, and internal friction angle. Also, the geometry of a typical type of instability feature, an elongate slide, is examined in detail. A rheological model describing a Coulomb-viscous plastic in effective stress terms is proposed to describe certain features of the mass movement process.

The model is used to derive equations defining the initial failure of the slopes, the mass movement thickness and accelerations, the velocity of flow in the gullies, and the shape of the mud nose scarp in the toe area. The model indicates the importance of pore pressure in controlling sediment dynamics.

Results of the model compare favorably with the observed shapes of mud noses. Initial failures on low slopes (~0.5°) are attributed to pore pressures approaching geostatic values. Flow velocities are calculated to be several feet per second, based upon estimated sediment viscosities and pore pressures during flow.

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ABSTRACT

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